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**The effects of anthropogenic greenhouse gases and aerosols on the inter-decadal change
of the South China Sea summer monsoon in the late 20th Century**

Zhongxi Lin¹ Buwen Dong² Zhiping Wen³

**1 Center for Monsoon and Environment Research/School of Atmospheric Sciences, Sun Yat-sen
University, Guangzhou, China**

**2 Department of Meteorology, National Centre for Atmospheric Science, University of Reading,
Reading, UK**

**3 Department of Atmospheric and Oceanic Sciences & Institute of Atmospheric Sciences, Fudan
University, Shanghai, China**

Corresponding author:

Zhongxi Lin,

Center for Monsoon and Environment Research/ School of Atmospheric Sciences

Sun Yat-Sen University,

Zhuhai, 519082 China

E-mail: linzhxi@mail2.sysu.edu.cn

Abstract: Analysis of observational precipitation indicates that in last few decades, the precipitation in boreal summer (June-August) over the South China Sea (SCS) exhibited an interdecadal variation, characterized by a decrease of 0.59mm/day from the period 1964-1981 to the period 1994-2011. Accompanied this decrease in precipitation is weakened monsoon circulation featured by an anti-cyclonic circulation anomaly over the SCS in the later period relative to the early period. This work investigates impacts of anthropogenic forcing changes on this interdecadal change in observations, quantify the relative roles of greenhouse gases (GHG) forcing and anthropogenic aerosol (AA) forcing. A set of experiments is designed using the atmospheric component of a state-of-the-art climate model coupled to a multi-level mixed-layer ocean model forced with GHG concentrations and AA emissions in two periods. Modeling results indicate a dominant role of anthropogenic forcing on the observed interdecadal precipitation decrease and weakened monsoon circulation over the SCS in the late 20th century in which AA forcing plays a more important role compared with GHG forcing. The mechanisms of GHG influences and AA induced changes are revealed by individual forcing experiments. Increasing GHG concentrations can suppress convection over the SCS summer monsoon (SCSSM) region by warming the tropical Pacific with an El-Niño like sea surface temperature (SST) pattern, which is associated with a weakened Walker circulation. The changes in AA emissions, mainly through increases in emissions over Asia, lead to cool SST in the north Indian Ocean and the western North Pacific (WNP), and result in changes in meridional SST gradient over the tropical Indian Ocean and the WNP in pre-monsoon seasons. This anomalous meridional SST gradient leads to anomalous local Hadley circulation, characterized by anomalous ascents around the equator and descents over monsoon region, which

49 suppresses convection over the SCS and reduces local precipitation.

50 **Key words:** South China Sea summer monsoon, anthropogenic greenhouse gases,
51 anthropogenic aerosols, precipitation.

52

53 1 Introduction

54 The South China Sea (SCS) locates in Southeast Asia between the equator and South
55 China coast, from 110°E to 120°E. It is in the center of the Asian-Australian monsoon system
56 and has a close relation with the East Asian monsoon, South Asian monsoon and Western
57 North Pacific monsoon due to geographic location (Tao and Chen 1987; Ding 1992,
58 Murakami and Matsumoto 1994; Wang 1994; Lau and Yang 1997). The climate over the
59 SCS is mainly controlled by monsoon circulation, characterized by a warm and wet westerly
60 low-level flow during boreal summer, and cold and dry easterly low-level flow during winter.
61 In summer, westerly flow comes from the Indian Ocean, turns into southerly and transports
62 moisture to the East Asian monsoon region. Over the SCS, this cyclonic circulation forms a
63 local monsoon trough and leads to convection and abundant precipitation.

64 The SCS summer monsoon (SCSSM) changes in multiple time-scales (Zhou et al. 2005;
65 Wang et al. 2009; Tong et al. 2009) and can be affected by many different factors, including
66 solar radiation on orbital time scale (Prell 2011), interdecadal time scale, and sea surface
67 temperature (SST) variability and land-sea thermal contrast on interannual time scale (Wang
68 and Zhang 2002; Lau et al. 2004; Zhou et al. 2005; Lau and Wang 2006; Yuan et al. 2008;
69 Wang and Qian 2009; Zhu et al. 2011). Its intensity is sensitive to the definition since the
70 SCSSM can be affected by the multiple monsoon subsystems, including the East Asian
71 summer monsoon (EASM), South Asian summer monsoon (SASM), and western North

72 Pacific summer monsoon (WNPSM). Previous researches use some indices to describe
73 characteristics of the SCSSM, including its southwesterly low-level flow, local convection
74 and moisture transport convergence (Lu and Chan 1999; Liang et al. 1999; Li and Zhang
75 1999; Dai et al. 2000; Liang et al. 1999; Wu and Liang 2001; Yao and Qian 2001; Zhang et
76 al. 2002; Wen et al. 2006; Wang et al. 2008). A meridional shear vorticity index (zonal wind
77 in the south SCS minus that in the north SCS; Wang et al. 2009) is created to describe the
78 strength of the SCSSM and is then widely used. Based on 7-year mean of this index, inter-
79 decadal changes are observed in the early 1970s and the early 1990s (Wang et al. 2009).
80 However, drivers and physical mechanisms for these interdecadal changes are not fully
81 investigated.

82 The interdecadal variation of the EASM and SASM has been attributed to both internal
83 variability associated with low frequency SST variability in the Atlantic and Pacific (i.e., the
84 Pacific Decadal Oscillation [PDO] and the Atlantic Multidecadal Oscillation [AMO], Goswami
85 et al. 2006; Lu et al. 2006; Zhou et al. 2009) and external forcing associated with changes
86 in greenhouse gases (GHG) and anthropogenic aerosol (AA) emissions (Song et al. 2014;
87 Li et al. 2015; Li and Ting 2017, Tian et al. 2018, Luo et al. 2019). Most of previous studies
88 agree that precipitation tends to increase in both EASM and SASM due to increased
89 moisture content in the atmosphere with increasing GHG forcings. Despite of the increased
90 moisture content, the effect of GHG and AA on the monsoon circulation is important and
91 requires further research.

92 The greenhouse gases (GHG) can change the SASM monsoon circulation by changing
93 the large-scale tropical Walker circulations and land-sea thermal contrast (Ueda et al. 2005).
94 In some model studies, the intensity of the Walker circulation connecting the tropical Pacific

95 and monsoon region tends to be weakened (Tanaka et al. 2004; Ueda et al. 2005, May 2002,
96 Vecchi and Soden 2006; Lau and Kim 2017), which suppresses the convection in the SASM
97 region. The land-sea thermal contrast is increased as the temperature over land response
98 to GHG increases is stronger than over ocean (Sutton et al. 2007; Dong et al. 2009). The
99 land-sea thermal contrast between South Asia and Indian ocean, which is an important
100 driver of the SASM (Wu et al. 2012), could lead to an enhanced circulation (Hu et al. 2000).
101 However, the weakened thermal contrast in the upper troposphere might play a more
102 important role in a long-term trend and weakened SASM monsoon circulation (Sun et al.
103 2010). For the EASM, Jiang and Wang (2005) and Song et al. (2014) suggested that the
104 summer monsoon circulation over East Asia is slightly enhanced based on model
105 ensembles. It is further revealed that the increased GHG leads to enhanced western North
106 Pacific Subtropical High (WNPSH) and increased southerly on the western boundary of
107 WNPSH (Tian et al. 2018). Over ocean, Lee et al. (2008) suggested that the low-level
108 convergence decreases over the western North Pacific and the WNPSM weakens despite
109 of increasing moisture in the atmosphere in response to CO₂ increases.

110 Some anthropogenic aerosols (AA) can cause surface cooling by scattering and
111 absorbing solar radiation and change the radiative properties of clouds (Boucher et al. 2013).
112 Meanwhile, black carbon and organic carbon can absorb reflected solar radiation to increase
113 the atmospheric temperature (Menon et al. 2002; Ming et al. 2010; Ramanathan et al. 2001).
114 As AA emissions increased during the last few decades in Asia, it might affect the adjacent
115 precipitation by altering the land-sea thermal contrast and relative monsoon circulation (Ming
116 and Ramaswamy 2011; Rotstayn and Lohmann 2002; Dong et al. 2016). In SASM region,
117 the monsoon circulation changes in response to increased AA emissions are characterized

118 by an anomalous meridional overturning circulation that tends to decrease precipitation over
119 the Indian peninsula (Bollasina et al. 2011). Some model studies suggest the resulted
120 anomalous meridional SST gradient over the Indian Ocean with cooling on the North Indian
121 Ocean in response to AA emission changes is the main driver (Ganguly et al. 2012; Ganguly
122 et al. 2012; Ramanathan et al. 2005; Chung et al. 2006; Luo et al. 2019). In EASM region,
123 observations show a wet trend in south China and a drought trend in north in past several
124 decades and this pattern is considered as a result of increased aerosol emissions in China
125 (Wang 2001; Ding et al. 2008, 2009; Menon et al. 2002; Xie et al. 2016; Tian et al. 2018).
126 Some studies suggest that the aerosol can cause a regional cooling over land and leads to
127 a weakened land-sea thermal contrast and thus weakened EASM circulation and
128 precipitation (Ye et al. 2013; Song et al. 2014, Dong et al. 2016).

129 In Asian monsoon region, GHG and AA do not always play similar roles for the monsoon
130 decadal variations. Lau and Kim (2017) suggested that increases in GHG since 1950s
131 generate an increased land-sea thermal contrast but it is masked by the cooling induced by
132 AA forcing. This study also suggested that the EASM and SASM are having different
133 sensitivity to GHG and AA changes. In EASM region, as the circulation is weakened by AA
134 effect while GHG's role is negligible due to the offset by its dynamic effect and
135 thermodynamic effect, AA forcing is more dominating in the decreased monsoon circulation
136 and tends to reduce the precipitation (Zhang and Li 2016; Song et al. 2014). Luo et al. (2019)
137 suggested that reduction in precipitation in South Asia in last few decades is related to the
138 weakened SASM circulation in response to changes in both GHG and AA and that
139 circulation change induced precipitation decrease is stronger than the increase related to
140 moisture increase in the atmosphere. All above mentioned studies are on changes of GHG

141 and AA emissions on either the EASM or SASM. In contrast, the anthropogenic influences
142 on the SCSSM and physical processes involved are not fully investigated.

143 This study aims to elucidate the processes of how GHG and AA affect the inter-decadal
144 change of the SCSSM in the late 20th century, focusing on: 1) What is the role of
145 anthropogenic forcing to the observed inter-decadal change? 2) What are the relative roles
146 of changes in GHG and AA forcings? 3) What are the physical processes? We have
147 performed a set of model experiments using the atmospheric component of the state-of-the-
148 art HadGEM3 global climate model coupled to a multi-level mixed-layer ocean model to
149 address these questions.

150 Lots of studies about impacts of anthropogenic forcings on the EASM and SASM use
151 full ocean Coupled General Circulation Models (CGCM) including CMIP5 models (Song et
152 al. 2014; Seo et al. 2013; Chen et al. 2016; Sharmila et al. 2015), but these models usually
153 have significant bias on sea surface temperature (SST) in mean state (Li et al. 2012; 2015;
154 Levine et al. 2013). As the Australian-Asian monsoon is sensitive to SST, this bias may
155 cause even larger bias on monsoon simulation. Thus, MetUM-GOML models use an ocean
156 mixed-layer model replacing the 3D CGCMs, which can reduce the calculation cost and
157 have a smaller SST bias by a prescribed flux corrections (Hirons et al. 2015, Dong et al.
158 2017, Tian et al. 2018). So, this work will use a set of experiments based on the second
159 generation of GOML models (MetUM-GOML2).

160 The structure of this paper is as follows: Section 2 analyses the observational data and
161 describes the inter-decadal change of the SCSSM in the late 20th century. Section 3
162 introduces the model and detailed design of model experiments. Section 4 evaluates the
163 relative contributions of different forcings to the inter-decadal change. Sections 5 and 6

164 reveal the physical processes of changing GHG or AA forcing on the SCSSM respectively.
165 Finally, our conclusions are drawn in Section 7.

166

167 **2 Observed inter-decadal change**

168 In this study, we use monthly NOAA's Precipitation Reconstruction (PREC) dataset
169 (Chen et al. 2002) and precipitation based on GPCP (Adler et al. 2003). Observed 3D
170 monthly horizontal velocity and vertical velocity come from NCEP Reanalysis 1 (NCEP1;
171 Kalnay et al. 1996) while monthly sea level pressure (SLP) is obtained from Hadley Centre
172 Sea Level Pressure dataset (HadSLP2; Allan and Ansell 2006). All variables are averaged
173 from June to August to describe the boreal summer mean.

174 To show the inter-decadal variability (IDV) of rainfall change in observations, the area
175 averaged precipitation in the South China Sea region (10°N-20°N, 105°E-120°E) in JJA is
176 defined as an index and its time evolution is shown in Figure 1a. One of the most important
177 features of the time evolution of this precipitation index is the inter-decadal variation across
178 the 1990s, characterized by a positive precipitation anomaly from 1964 to 1981 (early period,
179 EP) and a negative precipitation anomaly from 1994 to 2011 (present day, PD). These
180 periods are selected also to avoid the impact of volcanic eruptions. The mean anomaly in
181 these two periods is 0.45 mm/day in EP and -0.14 mm/day in PD, respectively, which means
182 a -0.59 mm/day (which is about 13.6% of the climatological mean of 1979-2018 in GPCP
183 dataset) decrease from EP to PD.

184 We compare the spatial pattern of precipitation and circulation associated with this inter-
185 decadal change (PD minus EP) to understand the physical processes involved. It is found
186 that the inter-decadal decrease in precipitation mainly centered in the SCS (up to 0.5-2.0

187 mm/day), and was accompanied by anomalous anticyclonic circulation in the lower
188 troposphere (fig. 1b) with anomalous easterlies across the SCS. The anomalous easterlies
189 over the SCS indicate weakened climatological SCS westerlies in the summer from EP to
190 PD. The anomalous anticyclonic circulation is also associated with anomalous divergence
191 in the lower troposphere and anomalous convergence in the upper troposphere (Fig. 1c)
192 with anomalous descent in the whole troposphere (Fig. 1d), indicating weakened convection
193 in the SCS and therefore leading to reduced precipitation.

194 What had caused this inter-decadal decrease in precipitation and monsoon circulation
195 over the SCS seen in observations? To find out drivers and understand the dynamical
196 mechanisms behind it, we perform a set of time slice experiments with the atmospheric
197 component of the state-of-the-art HadGEM3 global climate model coupled to a multi-level
198 mixed-layer ocean model with changes in GHG and AA emissions together from EP to PD
199 or changes in GHG and AA forcing separately.

200

201 **3 Models and experiments**

202 **3.1 MetUM-GOML2 model and experiments settings**

203 In this work, MetUM-GOML2 (e.g., Hirons et al. 2015) is used to perform a set of
204 experiments. The atmospheric component of the MetUM-GOML2 is the Met Office Unified
205 Model (MetUM) at the fixed scientific configuration Global Atmosphere 6.0 (GA6.0) with N96
206 horizontal resolution (1.875° longitude and 1.25° latitude) and 85 vertical levels. A detailed
207 description of GA6.0 is given by Walters et al. (2017). The atmospheric model includes an
208 interactive tropospheric chemistry scheme (eight aerosol species are included: ammonium
209 sulphate, mineral dust, fossil fuel black carbon, fossil-fuel organic carbon, biomass-burning,

210 ammonium nitrate, sea-salt, and secondary organic aerosols from biogenic emissions. Both
211 aerosol-radiation interaction and aerosol-cloud interaction (including aerosol indirect effect
212 and semi-indirect effect) are considered. More details about the parameterization of aerosol
213 indirect effect are described by Jones et al. (2011) and Bellouin et al. (2011, 2013).

214 The oceanic component is a Multi-Column K Profile Parameterization mixed-layer ocean
215 model. The horizontal resolution of MC-KPP is the same as AGCM where they are coupled,
216 and below each AGCM grid point in ocean region is one water column. The atmospheric
217 and oceanic components are coupled every 3 hours. Since MC-KPP only simulate vertical
218 mixing in each grid point and exclude the ocean dynamics, the climatological seasonal
219 ocean 3D flux corrections are applied on ocean temperature and salinity (Large et al. 1994).
220 These corrections are intended to represent the mean advection in ocean and account for
221 biases in atmospheric surface fluxes.

222 The set of experiments used in this study is the same as those used to investigate SASM
223 decadal changes in Luo et al. (2019). Here, we use greenhouse gases (GHG) and
224 anthropogenic aerosol (AA) forcings averaged over 1994-2011 as PD conditions. For EP
225 forcing, GHG forcing is an average of 1964-1981 while AA forcing is an average of 1970-
226 1981. The details of experiments are summarized in Table 1. Firstly, a relaxation experiment
227 is run for 12 years (E0) forced by PD GHG and AA forcing. In this experiment, the MC-KPP
228 profiles of temperature and salinity were relaxed to a PD climatological ocean temperature
229 and salinity from the Met Office ocean analysis data (Smith and Murphy 2007). Then we
230 diagnosed the seasonal cycle of climatological daily-mean 3D ocean temperature and
231 salinity corrections from E0. These temperature and salinity are the same and are imposed
232 in all free running coupled experiments. Therefore, response in the model experiments to a

particular forcing change includes both the fast response to the forcing change (i.e. surface and atmospheric process) and slow response mediated by forcing induced SST change. For free coupled experiments, the EP experiment (EP) is forced by GHG and AA forcing of EP while All forcing present day (PD-ALL) is forced by PD GHG and AA forcing. In addition, two extra experiments, present-day GHG forcing (PD-GHG) with PD GHG but EP AA and present-day AA forcing (PD-AA) with PD AA but EP GHG, that are used to analyze the individual effect of GHG and AA. All coupled experiments are run for 60 years and only last 55 years of each experiment are used for analysis. In all experiments, the annual mean concentrations of GHG are prescribed uniformly globally while the aerosol emissions include seasonal cycle and spatial distribution. The spatial distribution of changes in annual mean sulphur dioxide emissions between the two periods was illustrated in the Fig. 3 in Chen and Dong (2019) where changes are characterized by decreases over North America and Europe and increases over South Asia and East Asia.

The response of different forcing is estimated by the difference of climatology between a pair of experiments. The difference of PD-ALL and EP represent the combine effect of GHG and AA forcing difference, while PD-GHG minus EP and PD-AA minus EP are estimation of individual effect of GHG forcing and AA forcing. The statistical significance of the difference in a pair of models is assessed using a two tailed Student's t-test.

251

3.2 Model climatology

In this section, some climatological features of PD experiment are compared with observations. Fig.2a and b show the precipitation, SLP and 850hPa wind from June to August (JJA) in PD-ALL forcing experiment and Fig.2c and d show the SLP and 850hPa

256 wind from NCEP reanalysis 1 and observed precipitation from Global Precipitation
257 Climatology Project (GPCP; Adler et al. 2003). In observations, spatial pattern of SLP is
258 characterized by low SLP over the Eurasia continent and high SLP on the tropical Indian
259 ocean and western North Pacific (WNP). This SLP distribution is associated with westerly
260 monsoon flow from India and it turns into southwesterly with cyclonic circulation over the
261 SCS, known as the monsoon trough. This westerly monsoon flow transports a large amount
262 of water vapor to the SCS and leads to moisture convergence there associated with the
263 cyclonic flow, inducing convection and producing large precipitation in the region.

264 The general features of summer climatological precipitation, low-level wind and land-
265 sea contrast of SLP are well reproduced by GOML2 model. However, the westerly from the
266 Indian Ocean is stronger in the model due to the overestimated meridional land-sea SLP
267 contrast. Thus, the westerly wind and monsoon trough extend eastward into the WNP in the
268 model and the simulated precipitation in the monsoon trough is larger than in observations.

269 Despite these biases, the model simulates the summer climatological precipitation and
270 low-level circulation over East Asia and adjacent regions well compared with observations,
271 which suggests this model can be an appropriate tool to study the precipitation change over
272 the SCS in response to different forcings.

273

274 **4 Model simulated changes in response to different forcings**

275 To explore the effects of anthropogenic forcing on the SCSSM in model experiments,
276 the changes in precipitation, low-level circulation and 500hPa vertical velocity in response
277 to changes in both GHG and AA forcings are shown in Fig.3a. Model simulated responses
278 to changes in anthropogenic forcing from EP to PD are characterized by anomalous

279 easterlies from the WNP, across the SCS, into the Indian peninsula, indicating a weakening
280 of climatological westerlies. The weakened westerlies are associated with reduced
281 precipitation in latitude band of (10°N-20°N) and increased precipitation to the south. These
282 features in the model simulations are similar to observed inter-decadal changes between
283 PD and EP in observations (Fig.1b), indicating that the changes in anthropogenic forcing
284 might have played a dominant role in the observed inter-decadal changes of the SCSSM
285 and reduced precipitation over the SCS in the late 20th century.

286 Shown in Fig.3 b and c are model simulated changes in response to either changes in
287 GHG or AA forcing. The patterns of large-scale circulation and precipitation in response to
288 changes in AA forcing (Fig. 3c) are similar to changes in ALL forcing experiment (Fig. 3a).
289 Although precipitation under GHG forcing reduces as well, the easterly anomaly over the
290 SCS region is weak. These responses to individual forcing change indicate that model
291 simulated weakening of the SCSSM and reduced precipitation over the SCS from EP to PD
292 are predominantly due to changes in AA forcing (Fig.3a, c) with weak contributions from
293 changes in GHG (Fig. 3b).

294 To investigate the processes that contribute to the precipitation changes in the model
295 simulated responses to different forcings, the column integrated moisture flux transports and
296 their convergences are shown in Fig.4. The moisture flux transports and their convergences
297 are further decomposed into a dynamic component due to circulation change and a
298 thermodynamic component due to humidity change (e.g., Li and Ting 2017, Tian et al. 2018).

299 In many aspects, the patterns of changes in moisture flux convergences and their
300 magnitudes in response to different forcings (Fig. 4a, d, g) are similar to changes in model
301 simulated precipitation responses (Fig. 3). This indicates that it is the moisture flux

302 convergence responsible for model simulated precipitation changes. However, changes in
303 dynamic components and thermodynamics components show some different features in
304 response to different forcings. In response to ALL forcing change, the dynamic components
305 (Fig.4b) show anomalous westward moisture flux transport and anomalous moisture
306 transport divergence from the WNP, across the SCS, into the SASM region. These
307 anomalies in dynamic components are partially offset by anomalous eastward moisture
308 transports and convergences due to thermodynamic components related to warming mainly
309 induced by changes in GHG (this will be discussed in section 5). In response to changes in
310 GHG forcing, the model simulated moisture convergence changes (Fig.4d) are weak over
311 the SCS due to compensation between the dynamic component related to weakened
312 circulation (Fig.4e) and thermodynamic component related to increased water vapor in the
313 atmosphere (Fig. 4f). In response to changes in AA forcing, the changes in moisture flux
314 transports and their divergences are predominantly due to dynamic components related to
315 circulation changes with weak contribution from changes in humidity (Fig. 4g, h, and i).

316 To quantify the regional precipitation changes over the SCS, area averaged precipitation
317 changes in observations and model simulated changes in response to different forcings are
318 shown in Figure 5a. ALL forcing change leads to a decrease in precipitation by 0.32mm/day
319 and contributes approximately 72% of observed inter-decadal decrease (-0.44mm/day),
320 suggesting anthropogenic forcing changes from EP to PD might have been a dominating
321 effect for observed inter-decadal decrease in precipitation in the late 20th century. Individual
322 forcing both have a positive contribution on precipitation decrease with a dominant
323 contribution from changes in AA forcings. The regional averaged moisture transport
324 convergence is shown in Figure 5c. As the total moisture transport convergence shows a

325 similar change as precipitation in Fig.5a, the dynamic and thermodynamic components
326 clearly show how anthropogenic forcings change circulation and humidity, and contribute to
327 the interdecadal decreased precipitation. All forcing, GHG forcing and AA forcing result in
328 weakened dynamic moisture transport convergence. However, as GHG changes warm the
329 atmosphere significantly, the thermodynamic part of all forcing and GHG forcing increases
330 and this offsets the dynamic component partially. AA emission changes do not change the
331 humidity much, so the thermodynamic component in AA forcing is weak and the weakened
332 total moisture transport convergence is predominantly related to changes in circulation. It
333 should be noted a nonlinearity of changes in the SCSSM in response to changes in GHG
334 and AA emissions (i.e., the response to the combined forcings is not necessarily equal to
335 the sum of the responses to the individual forcings). In particular, the sum of precipitation
336 changes over the SCS in response to GHG and AA changes is larger than the change in
337 response to ALL forcing. The nonlinearity noticed in this study is consistent with some
338 previous studies (Feichter et al. 2004, Ming and Ramaswamy 2009, Shiogama et al. 2012).
339 However, detailed investigation of the causes in this nonlinearity is beyond the scope of this
340 study.

341 The regional averaged zonal wind components over southern SCS (5°N-15°N, 105°E-
342 120°E) are shown in Fig.5b. In observations and model experiments, zonal wind in this
343 region change most significantly. This figure further illustrates that the weakened
344 precipitation over the SCS in response to changes in anthropogenic forcing (ALL forcing) is
345 mainly contributed by the weakened zonal wind and reduced moisture transports, and the
346 response to AA forcing plays a dominant role. In summary, it has demonstrated that the
347 circulation and precipitation changes in response to changes in individual GHG or AA forcing

348 show different characteristics. Thus, in the next two sections, we are exploring physical
349 processes that are responsible for the large-scale circulation differences in response to
350 individual forcing respectively.

351

352 **5 Mechanism related to GHG forcing**

353 The model simulated precipitation changes under GHG forcing show some weak
354 decreases, mainly resulted from the moisture flux transport compensation between the
355 dynamic component related to weakened circulation and thermodynamic component related
356 to increased humidity. What processes are responsible for the circulation and humidity
357 changes in the model simulations?

358 In GHG forcing experiment, large scale anomalous descends cover a large area
359 including the SASM region and the SCS region, accompanied by a low-level easterly
360 anomaly (Fig.3b). These anomalous local circulation anomalies and descents are
361 associated with anomalous large-scale divergent circulations in the tropical Indian and
362 Pacific oceans (Fig. 6). These anomalous large-scale circulations indicate anomalous
363 convergence over the tropical eastern Pacific in the lower troposphere and anomalous
364 divergence in the upper troposphere with opposite changes over the tropical Indian ocean
365 and the western tropical Pacific, being associated with weakened Walker circulation.
366 Associated with these large-scale divergent circulations are that SST warming is stronger in
367 the eastern tropical Pacific (over 0.5°C) than western tropical Pacific (about 0.3°C) (Fig.7b),
368 indicating a weakening zonal SST gradient in the tropical Pacific. This association between
369 the change in zonal SST gradient and large scale divergent circulations is consistent with
370 previous researches which suggested that the zonal SSTA gradient change associated with

371 El-Nino like SST anomalies enhances convection over the eastern equatorial Pacific and
372 leads to a weakened Walker circulation over the tropics (Ju and Slingo 1995; Timmermann
373 et al. 1999; Wilhelm, 2002).

374 How do changes in GHG forcing lead to enhanced warming over the eastern tropical
375 Pacific than in western tropical Pacific? Firstly, GHG forcing changes surface radiation
376 balance. With increase in GHG forcing, both clear sky surface longwave radiation (clear-sky
377 LW) and surface longwave cloud radiative effect (CRE-LW) increase over the equatorial
378 Pacific (Fig.7c and 7d). Because mixed layer depth is deeper in the western tropical Pacific
379 than in the eastern tropical Pacific, the response of SSTA is weaker in the west and stronger
380 in the east. This zonal asymmetry leads to a weakened zonal SST gradient and Walker
381 circulation through the dynamic processes elucidated above. This mechanism agrees with
382 previous research (DiNezio et al. 2009; Yeh et al. 2009; Collins et al. 2010) which revealed
383 the relationship of climate change, El-Nino like SSTA change and relative trade wind change.
384 Secondly, as air temperature increases due to increasing GHG concentrations, atmosphere
385 can hold more moisture over the SCS. Warmer sea surface temperature can also increase
386 the evaporation locally so both these changes lead to increase in the water vapor in the
387 atmosphere (Fig.7a).

388 The results above indicate that the decreasing precipitation and weakened monsoon
389 circulation over the SCS in GHG forcing experiments are caused by anomalous descend
390 which is related to weakened Walker circulation.

391

392 **6 Mechanism related to AA forcing**

393 In response to changes in AA forcing, the changes in moisture flux transports and their

394 divergences are predominantly due to dynamic components related to circulation changes
395 with weak contribution from changes in humidity (Fig. 4g, h, i). In AA forcing experiment,
396 there are significant easterly anomalies over the monsoon region from the Arabian Sea to
397 the western North Pacific (Fig. 3c).

398 To understand the mechanism of these circulation changes, we analyze the zonal mean
399 (60°E - 150°E) meridional overturning circulation changes in response to AA forcing (Fig.8a).
400 These anomalous overturning circulations are characterized by anomalous descents around
401 monsoon region (10°N - 25°N) and anomalous ascends around the equator, indicating a
402 change in local Hadley circulation. Previous researches suggest that this anomalous local
403 Hadley circulation might be a result of SSTA meridional gradient (Ganguly et al. 2012;
404 Ganguly et al. 2012; Ramanathan et al. 2005; Chung et al. 2006; Luo et al. 2019). However,
405 the SSTA gradient change in JJA is weak in response to AA forcing (Fig. 8b). The SSTA
406 over ascend region is about -0.05°C and it is about 0°C over descend region (Fig8b). This
407 association between anomalous ascents (descents) with anomalous cold (weak) SST
408 seems counterintuitive. However, this simultaneous association includes atmospheric
409 feedbacks on SST changes. Thus, we explore the SSTA in pre-monsoon seasons (Fig.9).
410 There is no significant SSTA over the northern Indian Ocean (NIO) and the SCS in JJA, but
411 during December to February (DJF) and March to May (MAM), SST is cooled to the north
412 of the equator in this region (about -0.1°C to -0.3°C). This meridional SST gradient in pre-
413 monsoon seasons is key processes for local Hadley circulation anomaly in JJA.

414 To understand the process of how aerosol changes affect SSTA and circulation in the
415 pre-monsoon seasons, seasonal evolutions of zonal averaged changes over the monsoon
416 region (60°E - 150°E) are analyzed and they are illustrated in Fig. 10. Results indicate that

417 large summer monsoon precipitation and circulation changes occur in June and become
418 weaker in July and August (Fig.10a and 10b). The meridional SSTA gradient with relative
419 large cold anomalies of about 0.2°C in the northern subtropics sustains during pre-monsoon
420 seasons until it gets weaker after May and June when monsoon establishes (Fig.10c). This
421 SSTA pattern and associated meridional SST gradient (cooler in north than in south)
422 sustains until May and June when it drives a local Hadley circulation which causes an
423 anomalous descend over monsoon region and then weakened westerly low-level flow and
424 less precipitation.

425 This SSTA annual cycle is driven by surface radiation changes. In response to AA
426 forcing, clear-sky surface shortwave radiation (clear-sky SW) decreases all year round over
427 the ocean with maximum changes during SON and MAM (Fig.10d), which cause a cooler
428 SSTA locally. The clear-sky SW changes are closely related to the aerosol-radiation
429 interaction and aerosol-cloud interaction. Changes in aerosol optical depth (AOD), cloud
430 droplet effective radius (CDER) and cloud droplet number (CDN) are larger over the NIO
431 and the SCS in pre-monsoon seasons (largest during SON) than in JJA since aerosol
432 emissions are advected by the prevailing winds to these regions (Fig.11j-11i). These
433 changes are characterized by increases in AOD, decreases in CDER, and increases in CDN.
434 Note that seasonal changes in CDER show some differences in comparison with changes
435 in AOD and CDN since CDER-CDN relationship is not linear (Jones et al. 2001) and CDER
436 change is more sensitive to the basic state environment than CDN over ocean where the
437 atmosphere is cleaner (Carslaw et al. 2013). These changes in AOD, CDN, and CDER are
438 responsible for the seasonal evolutions of changes in surface clear sky SW and SST (Fig.
439 10c, d). As a result, the SSTA is cool over the NIO to equatorial IO since SON and sustains

440 until May and Jun.

441 In May and June, the SST is still cooler over the NIO than around the equator, which
442 drives a local Hadley circulation. However, upward surface latent heat flux reduces related
443 to weakened low-level westerlies (anomalous easterlies) over monsoon region (positive
444 downward in Fig.10e). Meanwhile shortwave cloud radiative effect shows positive changes
445 (CRE-SW, Fig.10f) due to less cloud cover associated with reduced convection. These
446 feedbacks associated with the changes in circulation and surface radiation damp cold SST
447 anomalies induced by AA change in pre-monsoon seasons, leading to weak SST anomalies
448 in JJA and weak meridional SST gradient.

449 In summary, when changes in AA emissions are advected by prevailing winds from the
450 Asian continent to the NIO and the SCS in pre-monsoon seasons, the SST is cooled by
451 reduced clear sky SW while changes in both clear sky SW radiation and SST are weak in
452 the southern hemisphere. These result in anomalous meridional SST gradient in Asian
453 sector and this meridional SST gradient sustains until boreal summer, which in turn drives
454 weakened monsoon circulation and precipitation over the SCS in summer. In May and June
455 when precipitation is reduced, the SST warms over the NIO due to increased shortwave
456 cloud radiative effect (Fig. 10f) related to reduced cloud cover and weakened latent heat flux
457 (Fig. 10e) related to weakened westerlies. These processes tend to damp cold SST
458 anomalies in pre-monsoon seasons lead to weak summer mean SST anomalies over the
459 NIO (Fig.9d, Fig. 10c).

460

461 7 Conclusion and discussion

462 Base on observations, there is a significant inter-decadal decrease of precipitation over

463 the SCS between 1964-1981 and 1994-2011 during boreal summer monsoon season (JJA).
464 Corresponding to the change in precipitation is a weakened SCSSM circulation,
465 characterized by anomalous descend over the SCS, weaker westerly monsoon flow in the
466 lower troposphere and less moisture flux transport convergence to the SCSSM region. In
467 this study, a set of experiments based on MetUM-GOML2 model is used to investigate the
468 role of anthropogenic forcing (including GHG and AA emission) changes and quantify the
469 relative roles of GHG and AA forcing, on the inter-decadal decrease of the SCSSM and
470 precipitation. The main results are summarized as follows:

471 1. Model results suggest that the changes in anthropogenic forcing from EP (1964-
472 1981) to PD (1994-2011) might have played a dominant role in the observed inter-decadal
473 changes of the SCSSM and reduced precipitation in the late 20th century in which AA forcing
474 plays a more important role compared with GHG forcing.

475 2. Increased GHG concentrations lead to positive changes in surface longwave
476 radiation over the tropical Pacific. Due to the asymmetric mixed layer depth, warming in the
477 eastern equatorial Pacific is greater than in the western tropical Pacific. This El-Niño-like
478 SSTA pattern leads to a weaker Walker circulation and causes a large-scale descend in
479 both Indian monsoon and SCS monsoon region, and causes a weaker westerly low-level
480 flow. This weakened monsoon circulation leads to weakened moisture flux transport
481 convergence and therefore reduced precipitation over the SCS. Meanwhile, the warm SST
482 anomalies lead to increases of evaporation and the water vapor in the atmosphere. This
483 leads to increased moisture flux transport convergence and therefore enhanced
484 precipitation over the SCS. Due to this compensation between dynamical moisture transport
485 and thermodynamic moisture transport, the changes in moisture transport convergence, and

486 therefore precipitation over the SCS are weak.

487 3. In response to changes in AA emissions, CDN and AOD increase while CDER
488 decreases over continental Asia due to increased AA emissions over Asia. In pre-monsoon
489 seasons, aerosol emissions are transported to the northern Indian Ocean and the SCS by
490 prevailing winds, leading to increases in CDN and AOD and decreases in CDER in these
491 regions. These changes lead to reduced surface clear-sky shortwave radiation in the
492 northern subtropics with weak changes in southern hemisphere. As a result, the SST in the
493 northern subtropics decreases by about 0.2°C in pre-monsoon seasons and this cold
494 anomaly sustains until the establishment of summer monsoon. This nonuniform SST change
495 leads to a meridional SST gradient which in turn induces a local Hadley circulation change
496 with anomalous descending at monsoon region and the anomalous easterly wind in the
497 lower troposphere which transport less moisture to the SCS monsoon region, leading to
498 reduced precipitation.

499 Our modelling results suggest that changes in GHG concentrations and AA emissions
500 might have played an important role in decreased SCSSM and reduced precipitation on the
501 SCS in the late 20th century in which AA forcing plays a more important role compared with
502 GHG forcing. This result is in agreement to some previous researches about role of AA
503 changes in the East Asian summer monsoon based on CMIP5 models (Song et al. 2014;
504 Zhang and Li 2016) .

505 Although the observed inter-decadal decrease of SCSSM precipitation can be captured
506 by this model, it should be pointed out that these results and processes are based on this
507 specific model and experiment design. For example, these experiments use GHG and AA
508 forcing of two specific periods (1964-1981 and 1994-2011) and the ocean model is a mixed

509 layer model. Therefore, the role of ocean dynamics are not included in our experiments. In
510 the future, the reliability of the physical processes in this study can be tested by comparing
511 with other models' results which consider the oceanic circulation changes in response to
512 different forcings. In addition, it should be noted a nonlinearity of changes in the SCSSM in
513 response to changes in GHG and AA emissions (i.e., the response to the combined forcings
514 is not necessarily equal to the sum of the responses to the individual forcings). The
515 nonlinearity noticed in this study is consistent with some previous studies (Feichter et al.
516 2004, Ming and Ramaswamy 2009, Shiogama et al. 2012), and needs to be investigated
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518

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530 Reference:

- 531 Adler RF, Huffman GJ, Chang A, et al (2003) The Version-2 Global Precipitation Climatology
532 Project (GPCP) Monthly Precipitation Analysis (1979–Present). J Hydrometeorol. doi:
533 10.1175/1525-7541(2003)004<1147:tvgps>2.0.co;2
- 534 Allan R, Ansell T (2006) A new globally complete monthly historical gridded mean sea level
535 pressure dataset (HadSLP2): 1850–2004. J Clim. doi: 10.1175/JCLI3937.1
- 536 Bellouin N, Mann GW, Woodhouse MT, et al (2013) Impact of the modal aerosol scheme
537 GLOMAP-mode on aerosol forcing in the hadley centre global environmental model.
538 Atmos Chem Phys. doi: 10.5194/acp-13-3027-2013
- 539 Bellouin N, Rae J, Jones A, et al (2011) Aerosol forcing in the Climate Model
540 Intercomparison Project (CMIP5) simulations by HadGEM2-ES and the role of
541 ammonium nitrate. J Geophys Res Atmos. doi: 10.1029/2011JD016074
- 542 Bollasina MA, Ming Y, Ramaswamy V (2011) Anthropogenic Aerosols and the Summer
543 Monsoon. Science (80-) 334:502–505
- 544 Boucher O, Randall D, Artaxo P, Bretherton C, Feingold G, Forster P, Zhang X (2013)
545 Clouds and aerosols. In: Stocker T et al (eds) Climate change 2013: the physical
546 science basis. Contribution of working group I to the fifth assessment report of the
547 intergovernmental panel on climate change. Cambridge University Press, Cambridge,
548 pp 571–657
- 549 Carslaw KS, Lee LA, Reddington CL, Pringle KJ, Rap A, Forster PM, Mann GW, Spracklen
550 DV, Woodhouse MT, Regayre LA, Pierce JR (2013) Large contribution of natural
551 aerosols to uncertainty in indirect forcing. Nature 503:67–71
- 552 Charlson RJ, Schwartz SE, Hales JM, et al (1992) Climate forcing by anthropogenic
553 aerosols. Science (80-). doi: 10.1126/science.255.5043.423
- 554 Chen L and O. W. Frauenfeld (2016) A comprehensive evaluation of precipitation
555 simulations over China based on CMIP5 multimodel ensemble projections. J Geophys

556 Res Atmos 121:5430–5452. doi: 10.1002/2013JD021190.Received

557 Chen M, Xie P, Janowiak JE, Arkin PA (2002) Global Land Precipitation: A 50-yr Monthly
 558 Analysis Based on Gauge Observations. J Hydrometeorol. doi: 10.1175/1525-
 559 7541(2002)003<0249:glpaym>2.0.co;2

560 Chen W, and Dong B (2019) Anthropogenic impacts on recent decadal change in
 561 temperature extremes over China: Relative roles of greenhouse gases and
 562 anthropogenic aerosols. *Climate Dyn.* 52, 3643–3660, doi: 10.1007/s00382-018-4342-
 563 9.

564 Chung CE, Ramanathan V (2006) Weakening of north Indian SST gradients and the
 565 monsoon rainfall in India and the Sahel. J Clim. doi: 10.1175/JCLI3820.1

566 Collins M, An S II, Cai W, et al (2010) The impact of global warming on the tropical Pacific
 567 Ocean and El Niño. Nat Geosci 3:391–397. doi: 10.1038/ngeo868

568 Dai N, Xie A, Zhang Y (2000) Interannual and Interdecadal Variations of Summer Monsoon
 569 Activities over South China Sea[J]. Climatic and Environmental Research(in
 570 Chinese),5(4):363-374,doi:10.3878/j.issn.1006-9585.2000.04.04.

571 Ding Y (1992) Summer monsoon rainfall in China. J. Meteorol. Soc. Jpn. 70, 373–396

572 Ding Y, Sun Y, Wang Z, et al (2009) Inter-decadal variation of the summer precipitation in
 573 China and its association with decreasing Asian summer monsoon Part II: Possible
 574 causes. Int J Climatol 29:1926–1944. doi: 10.1002/joc.1759

575 Ding Y, Wang Z, Sun Y (2008) Inter-decadal variation of the summer precipitation in East
 576 China and its association with decreasing Asian summer monsoon. Part I: Observed
 577 evidences. Int J Climatol. doi: 10.1002/joc.1615

578 DiNezio PN, Clement AC, Vecchi GA, et al (2009) Climate response of the equatorial Pacific
 579 to global warming. J Clim 22:4873–4892. doi: 10.1175/2009JCLI2982.1

580 Dong B, Gregory JM, Sutton RT (2009) Understanding land-sea warming contrast in
 581 response to increasing greenhouse gases. Part I: Transient adjustment. J Clim

582 22:3079–3097. doi: 10.1175/2009JCLI2652.1

583 Dong B, Sutton RT, Highwood EJ, Wilcox LJ (2016) Preferred response of the East Asian
584 summer monsoon to local and non-local anthropogenic sulphur dioxide emissions. *Clim*
585 *Dyn* 46:1733–1751. doi: 10.1007/s00382-015-2671-5

586 Feichter J, Roeckner E, Lohmann U, Liepert B (2004) Nonlinear aspects of the climate
587 response to greenhouse gas and aerosol forcing. *J Clim* 17:2384–2398

588 Ganguly D, Rasch PJ, Wang H, Yoon JH (2012a) Fast and slow responses of the South
589 Asian monsoon system to anthropogenic aerosols. *Geophys Res Lett* 39:1–5. doi:
590 10.1029/2012GL053043

591 Ganguly D, Rasch PJ, Wang H, Yoon JH (2012b) Climate response of the South Asian
592 monsoon system to anthropogenic aerosols. *J Geophys Res Atmos* 117:1–20. doi:
593 10.1029/2012JD017508

594 Goswami B, Madhusoodanan M, Neema C, Sengupta D (2006) A physical mechanism for
595 North Atlantic SST influence on the Indian summer monsoon. *Geophys Res Lett* 33,
596 L02706. doi: 10.1029/2005GL024803

597 Guo L, Highwood EJ, Shaffrey LC, Turner AG (2013) The effect of regional changes in
598 anthropogenic aerosols on rainfall of the East Asian Summer Monsoon. *Atmos Chem*
599 *Phys*. doi: 10.5194/acp-13-1521-2013

600 Held IM, Soden BJ (2006) Robust responses of the hydrologic cycle to global warming. *J*
601 *Clim*. doi: 10.1175/JCLI3990.1

602 Hu ZZ, Latif M, Roeckner E, Bengtsson L (2000) Intensified Asian summer monsoon and its
603 variability in a coupled model forced by increasing greenhouse gas concentrations.
604 *Geophys Res Lett* 27:2681–2684. doi: 10.1029/2000GL011550

605 JIANG D and Wang H (2005) Natural interdecadal weakening of East Asian summer
606 monsoon in the late 20th century. *Chinese Sci Bull* 50:1923. doi: 10.1360/982005-36

607 Jones A, Roberts DL, Woodage MJ, Johnson CE (2001) Indirect sulphate aerosol forcing in

608 a climate model with an interactive sulphur cycle. *J Geophys Res* 106 D17: 20,293–
 609 20,310

610 Jones CD, Hughes JK, Bellouin N, et al (2011) The HadGEM2-ES implementation of CMIP5
 611 centennial simulations. *Geosci Model Dev.* doi: 10.5194/gmd-4-543-2011

612 Ju J, Slingo J (1995) The Asian summer monsoon and ENSO. *Q J R Meteorol Soc.* doi:
 613 10.1002/qj.49712152509

614 Kalnay E, Kanamitsu M, Kistler R, et al (1996) The NCEP/NCAR 40-year reanalysis project.
 615 *Bull Am Meteorol Soc.* doi: 10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2

616 Lamarque JF, Bond TC, Eyring V, et al (2010) Historical (1850-2000) gridded anthropogenic
 617 and biomass burning emissions of reactive gases and aerosols: Methodology and
 618 application. *Atmos Chem Phys.* doi: 10.5194/acp-10-7017-2010

619 Lau NC, Ann MJ, Wang H (2005). Simulations by a GFDL GCM of ENSO-related variability
 620 of the coupled atmosphere-ocean system in the East Asian Monsoon region. *East Asian*
 621 *Monsoon.*

622 Lau NC., Wang B. (2006) Interactions between the Asian monsoon and the El Niño/Southern
 623 Oscillation. In: *The Asian Monsoon.* Springer Praxis Books. Springer, Berlin, Heidelberg

624 Lau KM, Kim KM (2017) Competing influences of greenhouse warming and aerosols on
 625 Asian summer monsoon circulation and rainfall. *Asia-Pacific J Atmos Sci* 53:181–194.
 626 doi: 10.1007/s13143-017-0033-4

627 Lau KM, Yang S (1997) Climatology and interannual variability of the Southeast Asian
 628 summer monsoon. *Adv Atmos Sci.* doi: 10.1007/s00376-997-0016-y

629 Lee SM, Jhun JG, Kwon M, Kim W (2008) Change in the western North Pacific summer
 630 monsoon circulation due to the CO₂ increase in IPCC AR4 CGCMs. *Asia Pac J Atmos*
 631 *Sci* 44(4):351–368

632 Levine RC, Turner AG, Marathayil D, Martin GM (2013) The role of northern Arabian Sea
 633 surface temperature biases in CMIP5 model simulations and future projections of Indian

634 summer monsoon rainfall. *Clim Dyn* 41:155–172. doi: 10.1007/s00382-012-1656-x

635 Li C, and Zhang L, (1999) Summer monsoon activities in the South China Sea and its
636 impacts. *Chin. J. Atmos. Sci.*, 23, 257–266.

637 Li G, Xie SP (2012) Origins of tropical-wide SST biases in CMIP multi-model ensembles.
638 *Geophys Res Lett* 39:1–5. doi: 10.1029/2012GL053777

639 Li G, Xie SP, Du Y (2015) Monsoon-induced biases of climate models over the tropical
640 Indian Ocean. *J Clim* 28:3058–3072. doi: 10.1175/JCLI-D-14-00740.1

641 Li J, Wu Z, Jiang Z, He J (2010) Can global warming strengthen the East Asian summer
642 monsoon? *J Clim*. doi: 10.1175/2010JCLI3434.1

643 Li X, Ting M (2017) Understanding the Asian summer monsoon response to greenhouse
644 warming: the relative roles of direct radiative forcing and sea surface temperature
645 change. *Clim Dyn*. doi: 10.1007/s00382-016-3470-3

646 Li X, Ting M, Li C, Henderson N (2015) Mechanisms of Asian Summer Monsoon Changes
647 in Response to Anthropogenic Forcing in CMIP5 Models*. *J Clim*. doi: 10.1175/jcli-d-
648 14-00559.1

649 Li Z, Lau WKM, Ramanathan V, et al (2016) Aerosol and monsoon climate interactions over
650 Asia. *Rev. Geophys.*

651 Liang J, Wu S, and You J, (1999) The research on variations of onset time of the SCS
652 summer monsoon and its intensity. *Chin. J. Trop. Meteor.*, 15, 97–105.

653 Lu E. and Chan J, (1999) A Unified Monsoon Index for South China. *J. Climate*, 12, 2375–
654 2385, doi: 10.1175/1520-0442(1999)012<2375:AUMIFS>2.0.CO;2

655 Lu R, Dong B, Ding H (2006) Impact of the Atlantic Multidecadal Oscillation on the Asian
656 summer monsoon. *Geophys Res Lett* 33:1–5. doi: 10.1029/2006GL027655

657 Luo F, Dong B, Tian F, Li S (2019) Anthropogenically Forced Decadal Change of South
658 Asian Summer Monsoon Across the Mid-1990s. *J Geophys Res Atmos* 124:806–824.
659 doi: 10.1029/2018JD029195

660 Murakami T., Matsumoto J. (1994). Summer monsoon over the Asian continent and western
661 north Pacific. *J. Meteorol. Soc. Jpn.* 72, 719–745.

662 May W (2002) Simulated changes of the Indian summer monsoon under enhanced
663 greenhouse gas conditions in a global time-slice experiment. *Geophys Res Lett.* doi:
664 10.1029/2001GL013808

665 Menon S, Hansen J, Nazarenko L (2002) Climate Effects of Black Carbon Aerosols in China
666 and India. 297:2250–2254

667 Ming Y, Ramaswamy V (2009) Nonlinear climate and hydrological responses to aerosol
668 effects. *J Clim* 22:1329–1339

669 Ming Y, Ramaswamy V, Persad G (2010) Two opposing effects of absorbing aerosols on
670 global-mean precipitation. *Geophys Res Lett* 37:1–4. doi: 10.1029/2010GL042895

671 Ming Y, Ramaswamy V (2011) A model investigation of Aerosol-Induced changes in tropical
672 circulation. *J Clim.* doi: 10.1175/2011JCLI4108.1

673 Pitari G, Ackerman A, Adams P, et al (2001) Aerosols, their Direct and Indirect Effects. *Clim*
674 *Chang* 2001 Sci Basis Contrib Work Gr I to Third Assess Rep Intergov Panel Clim
675 *Chang* 5:289–348

676 Prell WL (2011) Variation of monsoonal upwelling: A response to changing solar radiation

677 Ramanathan V, Chung C, Kim D, et al (2005) Atmospheric brown clouds: Impacts on South
678 Asian climate and hydrological cycle. *Proc Natl Acad Sci.* doi:
679 10.1073/pnas.0500656102

680 Ramanathan V, Crutzen PJ, Kiehl JT, Rosenfeld D (2001) Atmosphere: Aerosols, climate,
681 and the hydrological cycle. *Science* (80-) 294:2119 – 2124. doi:
682 10.1126/science.1064034

683 Rotstayn LD, Lohmann U (2002) Tropical rainfall trends and the indirect aerosol effect. *J*
684 *Clim.* doi: 10.1175/1520-0442(2002)015<2103:TRTATI>2.0.CO;2

685 Seo KH, Ok J, Son JH, Cha DH (2013) Assessing future changes in the East Asian summer

monsoon using CMIP5 coupled models. J Clim 26:7662–7675. doi: 10.1175/JCLI-D-12-00694.1

Sharmila S, Joseph S, Sahai AK, et al (2015) Future projection of Indian summer monsoon variability under climate change scenario: An assessment from CMIP5 climate models. Glob Planet Change 124:62–78. doi: 10.1016/j.gloplacha.2014.11.004

Shiogama H, Stone DA, Nagashima T, Nozawa T, Emori S (2012) On the linear additivity of climate forcing-response relationships at global and continental scales Int. J Climatol 33:2542–2550

Smith DM, Murphy JM (2007) An objective ocean temperature and salinity analysis using covariances from a global climate model. J Geophys Res Ocean. doi: 10.1029/2005JC003172

Song F, Zhou T, Qian Y (2014) Responses of East Asian summer monsoon to natural and anthropogenic forcings in the 17 latest CMIP5 models. Geophys Res Lett. doi: 10.1002/2013GL058705

Sun Y, Ding Y, Dai A (2010) Changing links between South Asian summer monsoon circulation and tropospheric land-sea thermal contrasts under a warming scenario. Geophys Res Lett 37:1–5. doi: 10.1029/2009GL041662

Sutton RT, Dong B, Gregory JM (2007) Land/sea warming ratio in response to climate change: IPCC AR4 model results and comparison with observations. Geophys Res Lett 34:2–6. doi: 10.1029/2006GL028164

Tanaka HL, Ishizaki N, Kitoh A (2004) Trend and interannual variability of Walker, monsoon and Hadley circulations defined by velocity potential in the upper troposphere. Tellus, Ser A Dyn Meteorol Oceanogr 56:250–269. doi: 10.1111/j.1600-0870.2004.00049.x

Tao S, Chen L. (1987) A review of recent research on the East Asian summer monsoon in China. In: Chang, C.-P., Krishnamurti, T.N. (Eds.), Monsoon Meteorology. Oxford University Press, New York, pp. 60–92.

712 Tian F, Dong B, Robson J, Sutton R (2018) Forced decadal changes in the East Asian
713 summer monsoon: the roles of greenhouse gases and anthropogenic aerosols. *Clim*
714 *Dyn* 51:3699–3715. doi: 10.1007/s00382-018-4105-7

715 Timmermann A, Oberhuber J, Bacher A, et al (1999) Increased El Nino frequency in a
716 climate model forced by future greenhouse warming. *Nature*. doi: 10.1038/19505

717 Tong HW, Chan JCL, Zhou W (2009) The role of MJO and mid-latitude fronts in the South
718 China Sea summer monsoon onset. *Clim Dyn* 33:827–841. doi: 10.1007/s00382-008-
719 0490-7

720 Trenberth KE, Shea DJ (2006) Atlantic hurricanes and natural variability in 2005. *Geophys*
721 *Res Lett* 33:1–4. doi: 10.1029/2006GL026894

722 Ueda H, Iwai A, Kuwako K, Hori ME (2006) Impact of anthropogenic forcing on the Asian
723 summer monsoon as simulated by eight GCMs. *Geophys Res Lett* 33:20–23. doi:
724 10.1029/2005GL025336

725 Vecchi GA, Soden BJ (2007) Global warming and the weakening of the tropical circulation.
726 *J Clim*. doi: 10.1175/JCLI4258.1

727 Walters D, Boutle I, Brooks M, et al (2017) The Met Office Unified Model Global Atmosphere
728 6.0/6.1 and JULES Global Land 6.0/6.1 configurations. *Geosci Model Dev*. doi:
729 10.5194/gmd-10-1487-2017

730 Wang B, Huang F, Wu Z, et al (2009) Multi-scale climate variability of the South China Sea
731 monsoon: A review. *Dyn. Atmos. Ocean*.

732 Wang Z. & Qian Y. (2009) The relationship of land-ocean thermal anomaly difference with
733 mei-yu and South China Sea summer monsoon *Adv. Atmos. Sci.* 26: 169. doi:
734 10.1007/s00376-009-0169-y

735 Wang B, Yim SY, Lee JY, et al (2014) Future change of Asian-Australian monsoon under
736 RCP 4.5 anthropogenic warming scenario. *Clim Dyn*. doi: 10.1007/s00382-013-1769-x

737 Wang B, Zhang Q (2002) Pacific-East Asian teleconnection. Part II: How the Philippine Sea

738 anomalous anticyclone is established during El Niño development. J Clim. doi:
739 10.1175/1520-0442(2002)015<3252:PEATPI>2.0.CO;2

740 Wang C (2004) A modeling study on the climate impacts of black carbon aerosols. J
741 Geophys Res Atmos. doi: 10.1029/2003jd004084

742 Wen Z, Huang R, He H, and Lan G (2006) The influences of anomalous atmospheric
743 circulation over mid-high latitudes and the activities of 30-60d low frequency convection
744 over low latitudes on the onset of the South China Sea summer monsoon. Chin. J.
745 Atmos. Sci., 30, 952–964.

746 Wu G, Liu Y, He B, et al (2012) Thermal controls on the asian summer monsoon. Sci Rep
747 2:1–7. doi: 10.1038/srep00404

748 Wu S, and Liang J (2001) An index of South China Sea summer monsoon intensity and its
749 characters. Chin. J. Trop. Meteor., 17, 337–344.

750 Xie X, Wang H, Liu X, et al (2016), Distinct effects of anthropogenic aerosols on the East
751 Asian summer monsoon between multidecadal strong and weak monsoon stages, J.
752 Geophys. Res. Atmos., 121, 7026– 7040, doi:10.1002/2015JD024228.

753 Yao Y, and Qian Y (2001) A study on the South China Sea monsoon index and the
754 relationship between the index and regional rainfalls of China. J. Nanjing Uni., 37, 781–
755 788

756 Ye J, Li W, Li L, Zhang F (2013) “North drying and south wetting” summer precipitation
757 trend over China and its potential linkage with aerosol loading. Atmos Res 125–126:12–
758 19. doi: 10.1016/j.atmosres.2013.01.007

759 Yeh SW, Kug JS, Dewitte B, et al (2009) El Niño in a changing climate. Nature 461:511–514.
760 doi: 10.1038/nature08316

761 Zhang L, Li T (2016) Relative roles of anthropogenic aerosols and greenhouse gases in land
762 and oceanic monsoon changes during past 156 years in CMIP5 models. Geophys Res
763 Lett. doi: 10.1002/2016GL069282

764 Zhou T, Gong D, Li J, Li B (2009) Detecting and understanding the multi-decadal variability
765 of the East Asian Summer Monsoon – Recent progress and state of affairs, Meteorol.
766 Z., 18, 455–467.

767 Zhou W, Chan JCL (2005) Intraseasonal oscillations and the South China Sea summer
768 monsoon onset. Int J Climatol 25:1585–1609. doi: 10.1002/joc.1209

769 Zhou W, Chan JCL. ENSO and the South China Sea summer monsoon onset[J].
770 International Journal of Climatology, 2010, 27(2):157-167. doi : 10.1002/joc.1380

771 Zhu Y, Wang H, Zhou W, Ma J (2011) Recent changes in the summer precipitation pattern
772 in East China and the background circulation. Clim Dyn 36:1463 – 1473. doi:
773 10.1007/s00382-010-0852-9

774 Table 1 Summarize of numerical experiments

775

Experiment		Ocean	Radiative Forcing
E0	Relaxation run	Relax to PD mean 3D	Climatological PD (1994-2011)
		ocean temperature and	greenhouse gases (GHG) and
		salinity to diagnose	anthropogenic aerosol (AA) precursor
		climatological temperature	emissions. The GHG and AA forcing
EP	Early period experiment	and salinity flux correction	after 2006 are from RCP4.5 scenario
		climatological temperature	Climatological EP GHG (1964-1981)
		and salinity flux corrections	and EP AA(1970-1981) precursor
		from relaxation run	emissions
PD-ALL	Present day all forcing experiment	same as above	Climatological PD GHG and PD AA
			precursor emissions
PD-GHG	Present day GHG forcing	same as above	Climatological PD GHG and EP AA
			precursor emissions
PD-AA	Present day AA forcing	same as above	Climatological EP GHG and PD AA
			precursor emissions

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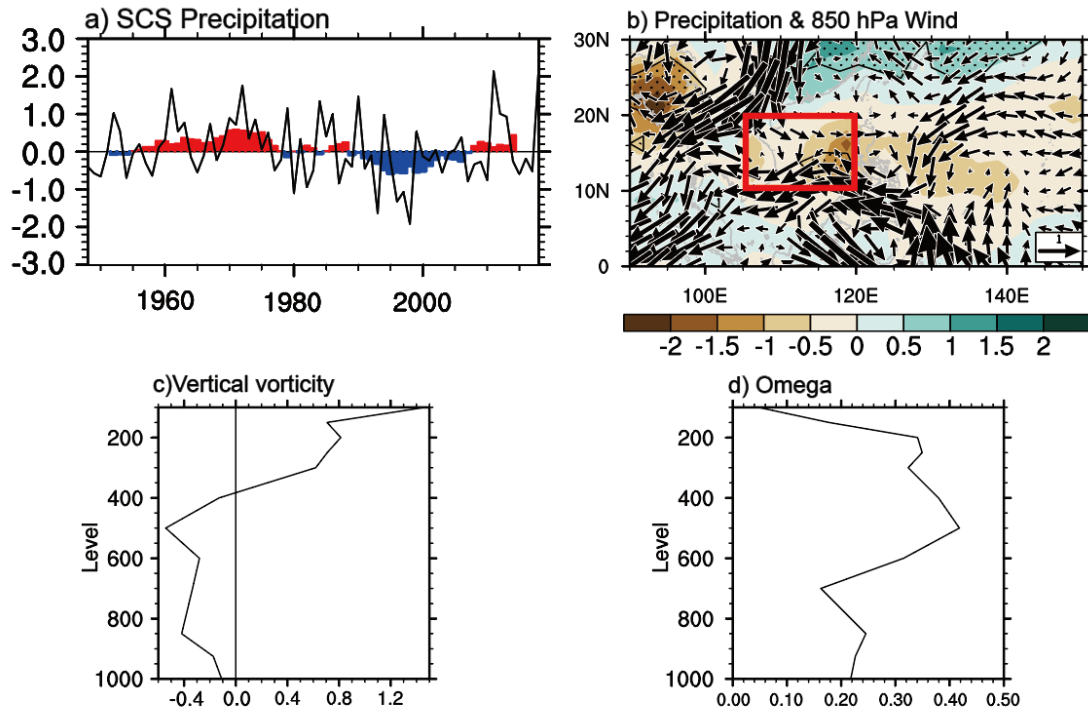


Fig.1 a) Time series (line) and 9 years moving mean (color bar) of (June-July-August) JJA mean precipitation (mm day^{-1}) over the South China Sea region (10°N - 20°N , 105°E - 120°E) based on NOAA_PREC data set. b)-d) changes between present day (1994-2011) and early period (1964-1981). b) Precipitation (mm/day) of NOAA_PREC and 850hPa wind(m/s) of NCEP reanalysis. Changes of precipitation in slash regions are statistically significant at the 10% level using a two-tailed Student's t-test. c) Vertical profile of vorticity ($10^{-6}/\text{s}$) over the SCS region, d) profile of vertical velocity (Omega , hPa/s) in pressure coordinate over the SCS. Red box highlights the SCS region.

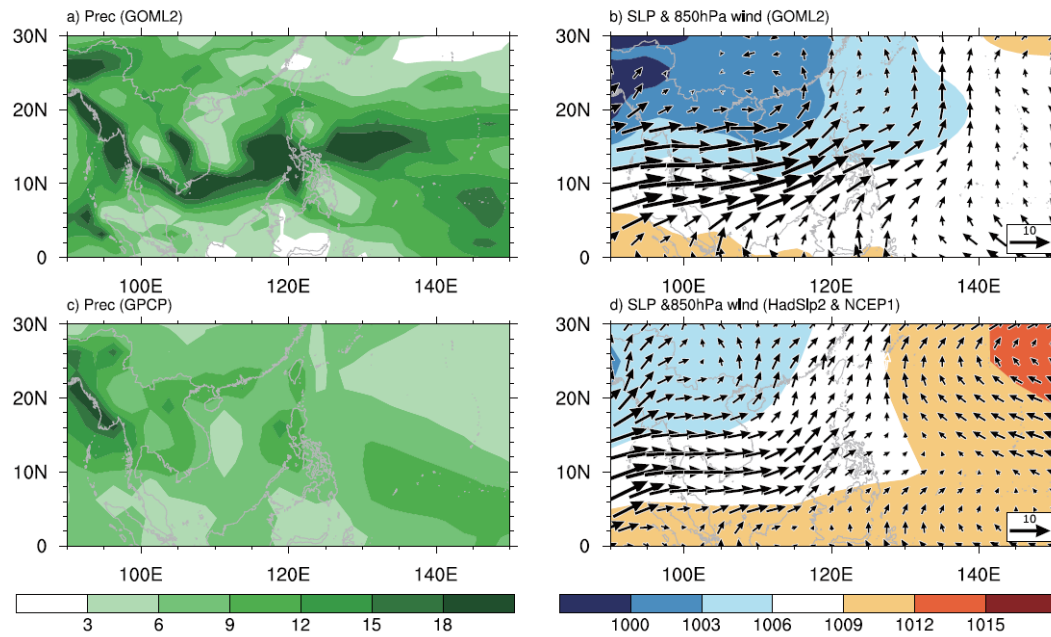
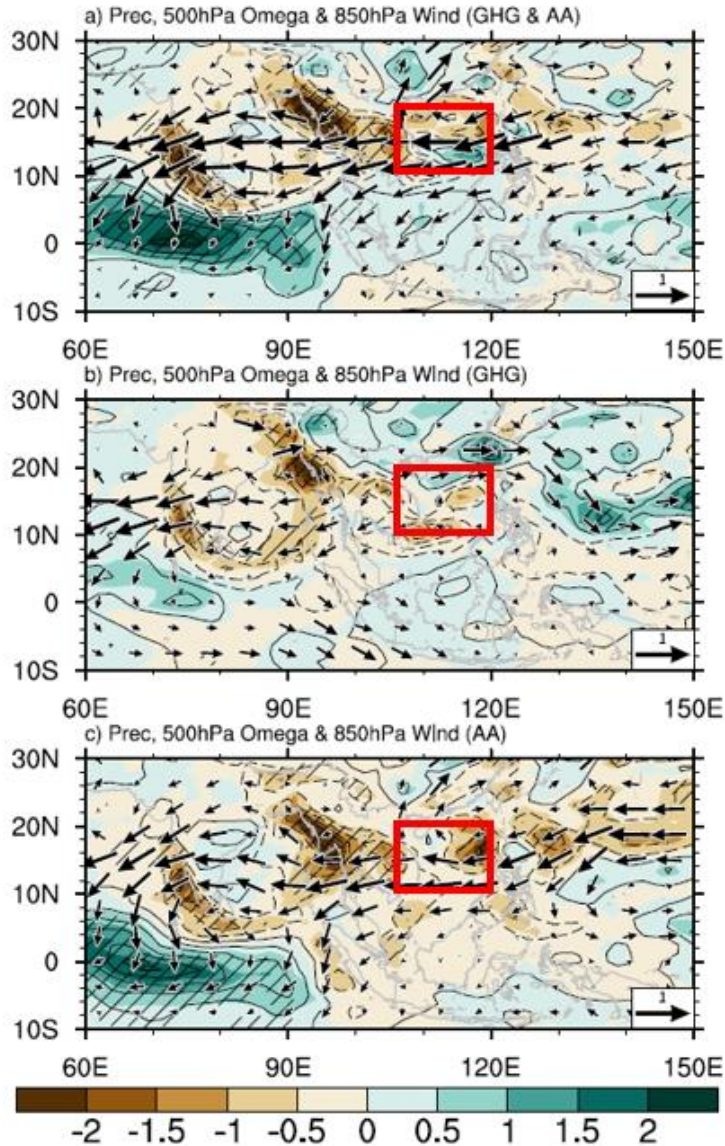


Fig.2 Present day (1994-2011) climatological precipitation (mm/day), sea level pressure (hPa), and 850hPa wind(m/s) in JJA. a) and b) GOML2 simulation. c) and d) GPCP precipitation, HadSLP2 SLP, and NCEP reanalysis 850 hPa wind.



794

795 Fig.3 Simulated changes of precipitation (shade, mm/day), 500hPa Omega (contour,

796 0.01Pa/s) and 850hPa wind (vector, m/s) in JJA. a) Response to anthropogenic forcing

797 (PD-ALL minus EP), b) response to GHG forcing (PD-GHG minus EP), and c) response to

798 AA forcing (PD-AA minus EP) Changes of precipitation in slash regions are statistically

799 significant at the 10% level using a two-tailed Student's t-test. Red box highlights the SCS

800 region.

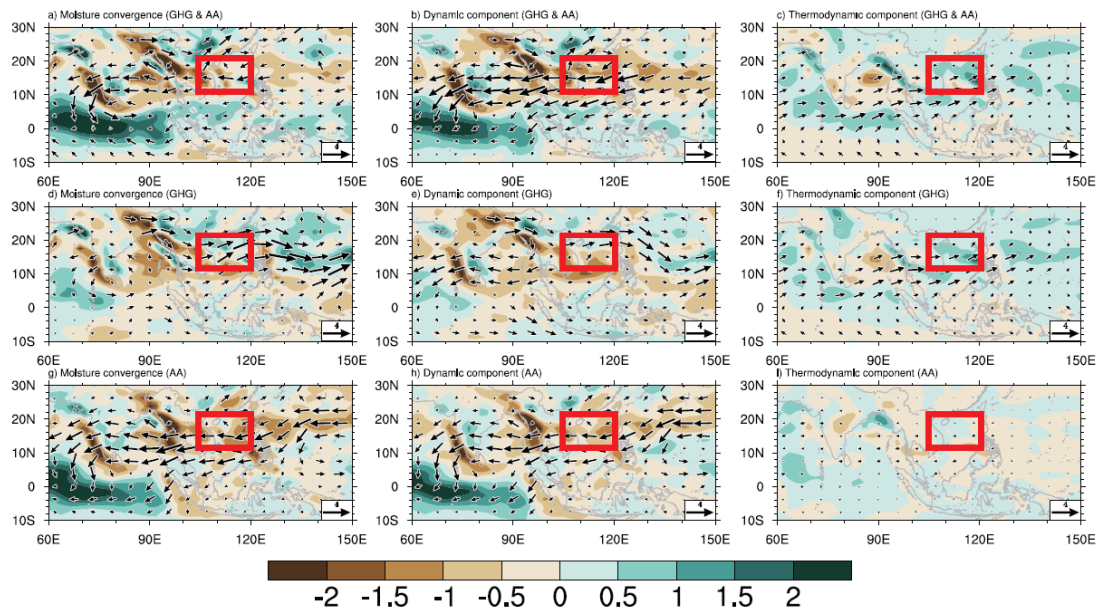
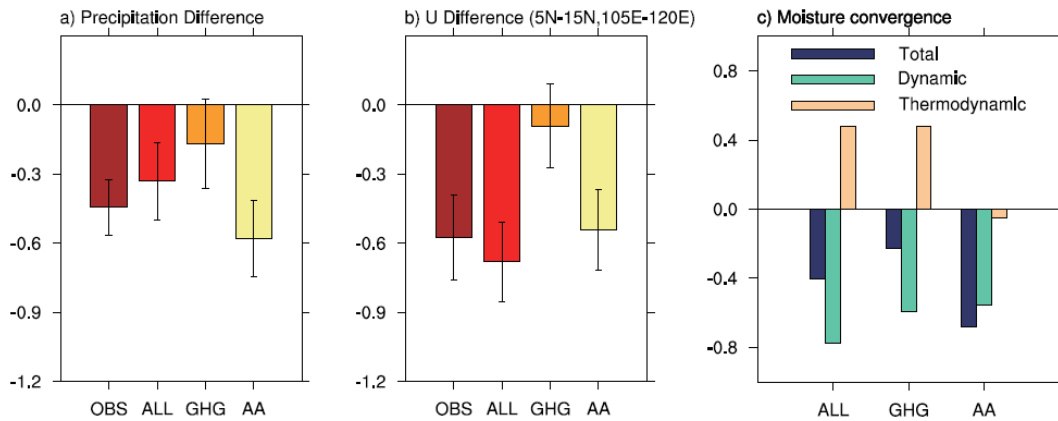


Fig.4. Simulated changes of vertically integrated moisture flux transport ($\text{kg/m}^2\text{s}$) and its convergence (mm/day) in JJA between PD and EP of All forcing (a, b, c), GHG forcing (d, e, f), and AA forcing (g, h, f). (a, d, g) total moisture flux and its convergence. (b, e, h) dynamic components related to changes in circulation. (c, f, i) thermodynamic components related to changes in humidity. Red box highlights the SCS region.

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811

812 Fig.5 a) precipitation (mm/day) change between PD and EP in observations
 813 (NOAA_PREC), all forcing, GHG forcing and AA forcing over the SCS region (10°N-
 814 20°N,105°E-120°E) in JJA.

815 b) zonal wind (m/s) change between PD and EP in observations (NCEP reanalysis 1), all
 816 forcing, GHG forcing and AA forcing over south SCS region (5°N-15°N,105°E-120°E) in
 817 JJA. The error bars indicate the standard errors of the mean difference.

818 c) simulated change of vertically integrated moisture transport convergence (blue, mm/day)
 819 between PD and EP in all forcing, GHG forcing and AA forcing over SCS region (10°N-
 820 20°N,105°E-120°E) in JJA. Green bars and yellow bars are the same as blue bars but for
 821 dynamic component and thermodynamic component of moisture convergence.

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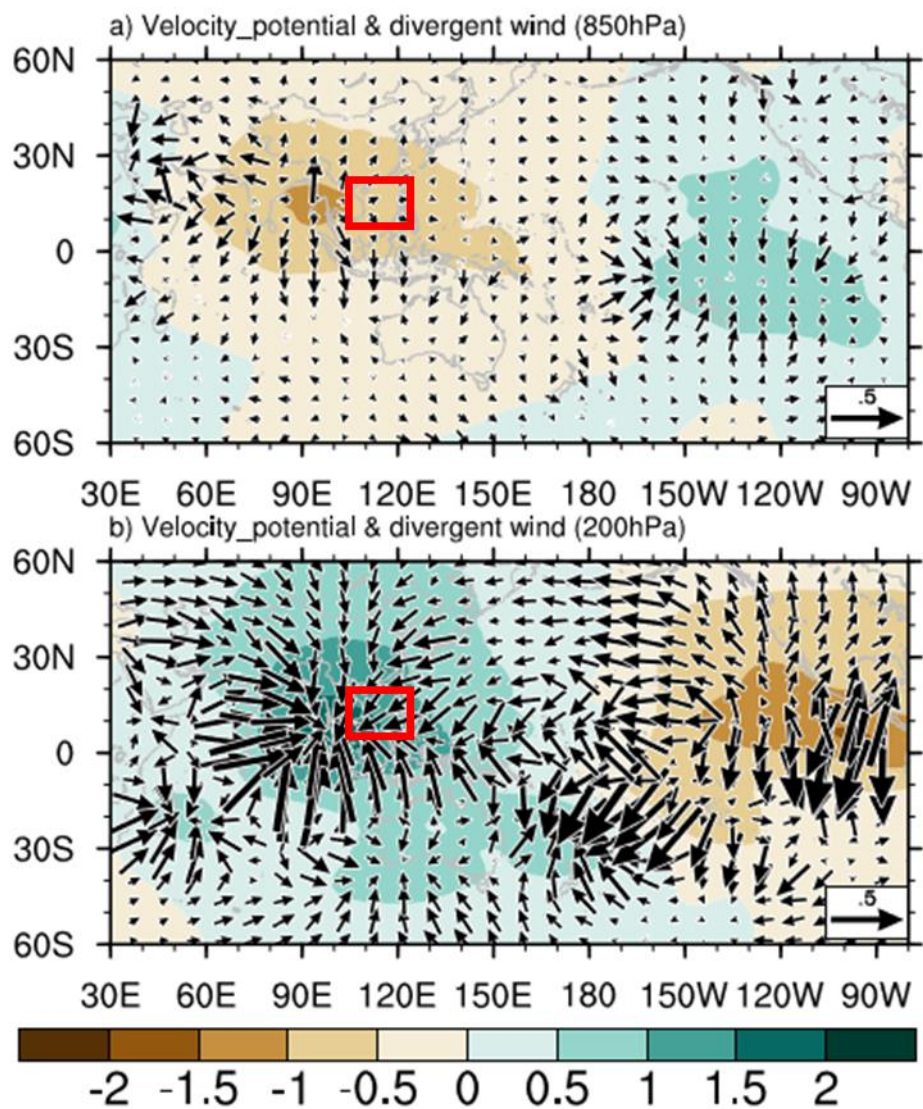


Fig.6 Simulated changes in velocity potential (shaded, $10^5 \text{ m}^2/\text{s}$) and divergent wind (vectors m/s) in JJA in response to GHG forcing (PD-GHG minus EP). a) at 850hPa, and b) at 200hPa. Red box highlights the SCS region.

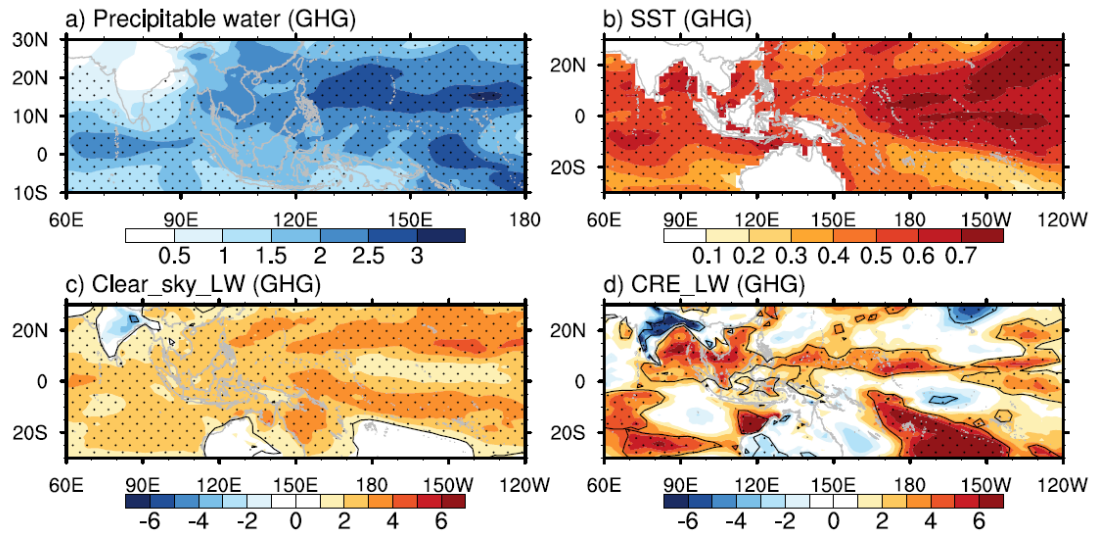
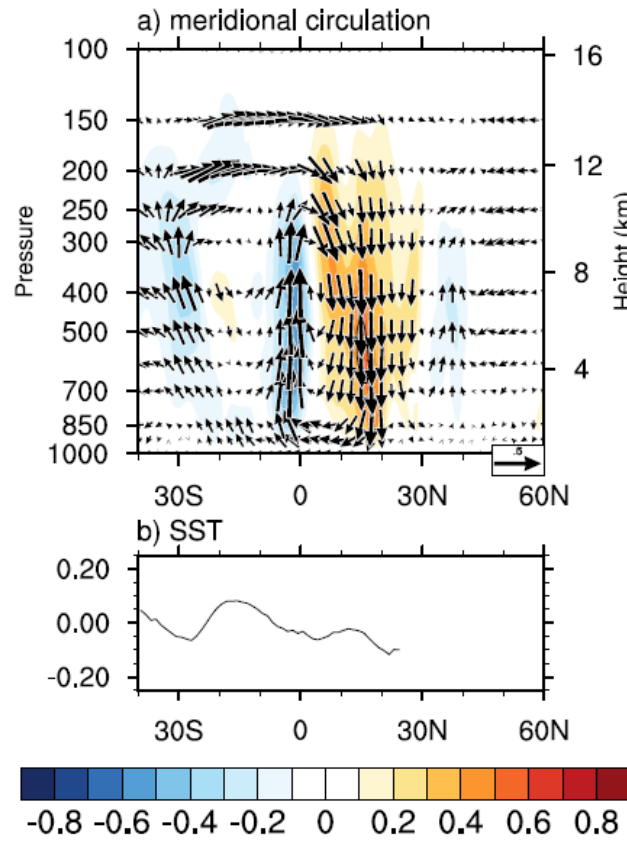


Fig.7 Simulated changes in JJA between PD and EP in response to GHG forcing. a) precipitable water (kg/m^2), b) sea surface temperature ($^{\circ}\text{C}$), c) surface clear sky net longwave radiation (W/m^2), and d) cloud radiative effect of longwave radiation (W/m^2). Radiations are positive downward. Dot regions are statistically significant at the 10% level using a two-tailed Student's t-test.

Zonal mean (60° E- 150° E) of JJA (AA)



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838 Fig.8 Simulated changes in JJA between PD and EP in response to AA forcing. a) zonal

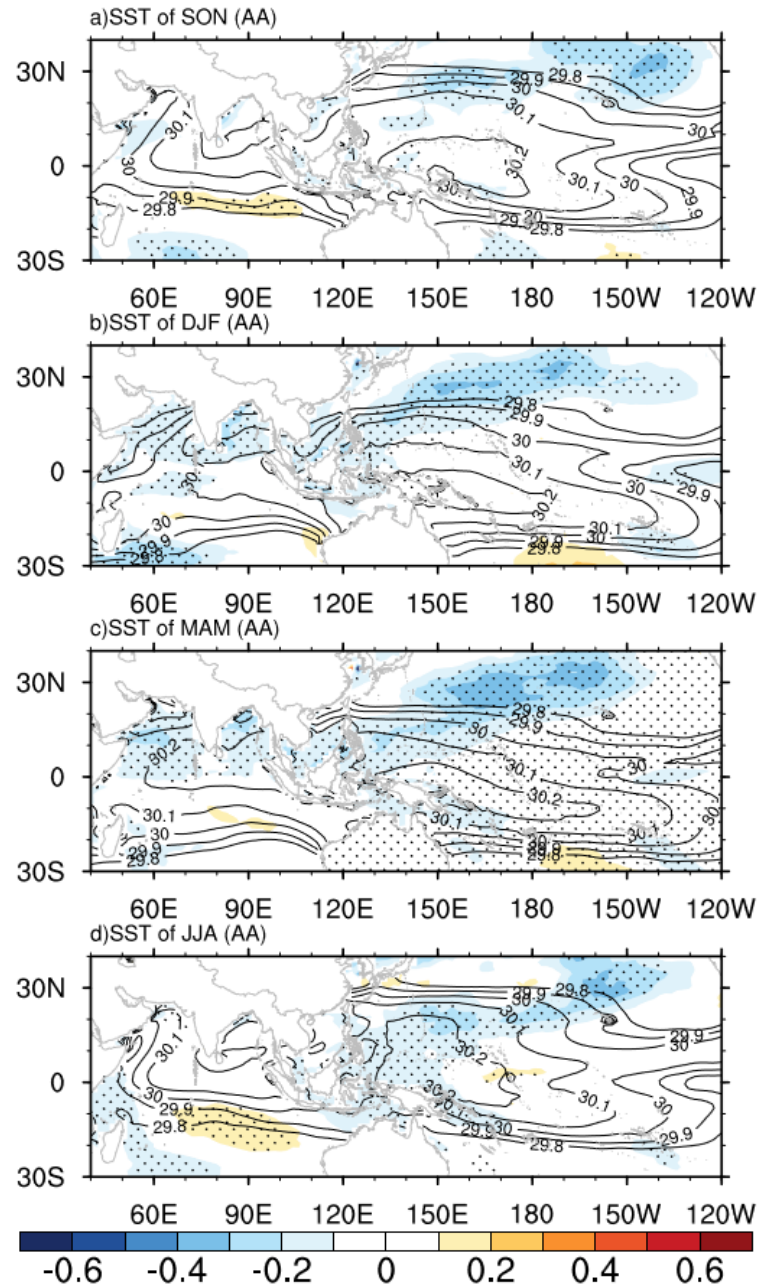
839 mean meridional circulation(vector) and vertical velocity(contour). Unit of meridional wind:

840 m/s, unit of vertical wind :0.01Pa/s.

841 b) zonal mean sea surface temperature ($^{\circ}$ C).

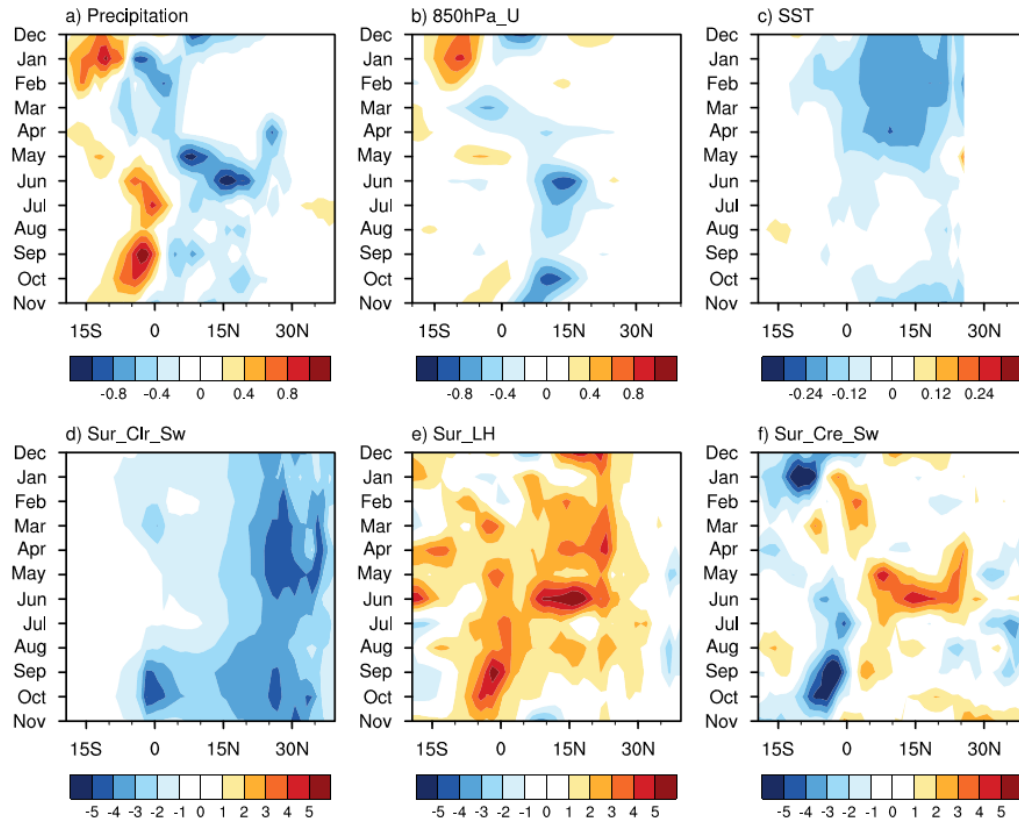
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843



844 Fig.9 Simulated seasonal changes (color, °C) in sea surface temperature between PD
 845 and EP in response to AA forcing and corresponding climatology (contour, °C) in the EP
 846 experiment. a) boreal autumn (SON), b) winter (DJF), c) spring (MAM), and d) JJA. Dot
 847 regions are statistically significant at the 10% level using a two-tailed Student's t-test.

Zonal mean of 60° E-150° E (AA)



848

849 Fig.10 Simulated seasonal evolutions of zonal averaged changes over the sector (60°E-

850 150°E) between PD and EP in response to AA forcing. a) precipitation change (mm/day)

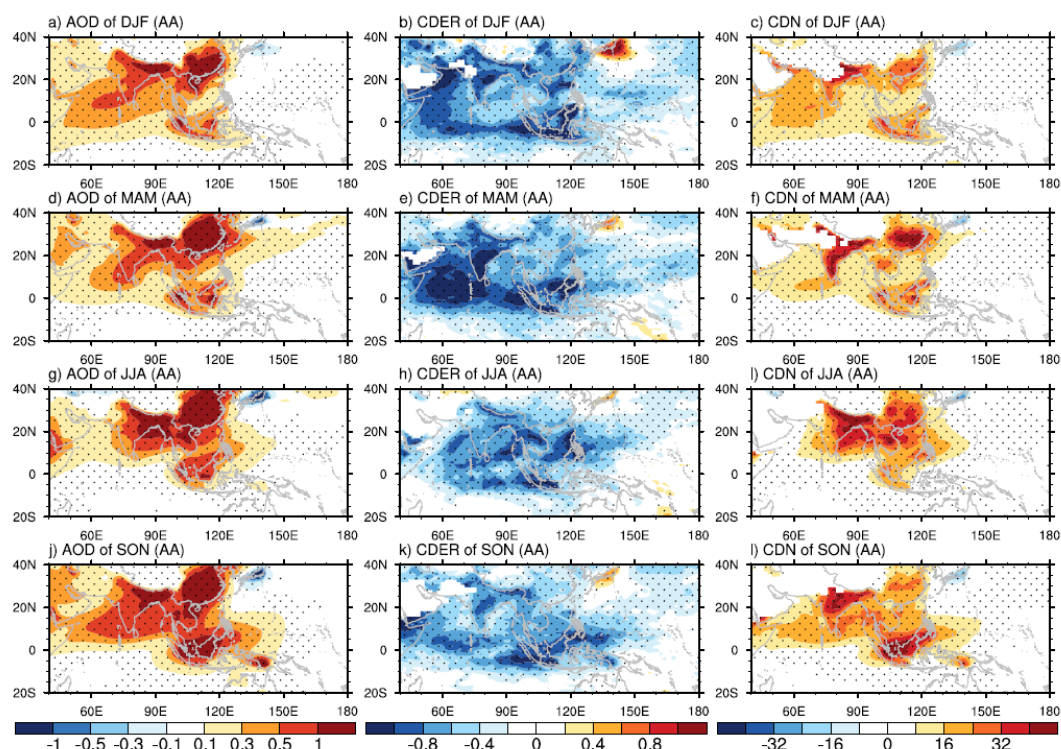
851 b)850hPa zonal wind(m/s), c) sea surface temperature(°C), d) surface clear sky net short

852 wave radiation, e) surface latent heat flux(W/m²), and f) surface cloud radiation effect of

853 short wave radiation (W/m²). Radiation and flux are positive downward.

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856

857 Fig.11 Simulated seasonal changes of aerosol optical depth (AOD, left panels, unit in 0.1)
 858 at 0.55 μm , cloud droplet effective radius (CDER, middle panels, unit in μm) and cloud
 859 droplet number (CDN, right panels, unit in 10^9 m^{-2}) between PD and EP in response to AA
 860 forcing. a)-c) DJF. d)-f) MAM, g)-h) JJA, j)-l) SON.

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